THE METEOROLOGICAL MAGAZINE

August 1987

Met.O.978 No. 1381 Vol. 116

METEOROLOGICAL OFFICE Met.O.978

THE METEOROLOGICAL MAGAZINE

No. 1381, August 1987, Vol. 116

551.509.313:551.521.14:551.551.8:551.588.3

An introduction to the parametrization of land-surface processes Part I. Radiation and turbulence*

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Summary

An introduction is given to the subgrid-scale land-surface processes which, it is generally acknowledged, need to be included by parametrization in three-dimensional numerical models for studying climate and climate change, and for numerical weather prediction. The discussion is restricted in the main to the relatively simple case of non-vegetated land surfaces.

Part I describes the general boundary conditions for momentum transfer and the balance equations for energy and mass (moisture) transfer at a bare-soil surface. Also the surface radiative properties and fluxes, and the physical character and the parametrization of the surface turbulent exchanges, are considered.

Part II (Carson 1987) will concentrate upon soil heat conduction and the land-surface temperature, and surface hydrology and the soil water budget.

1. Introduction

The atmospheric boundary layer is the lowest layer of the atmosphere characterized by significant vertical flux divergences of momentum, heat and moisture, which result directly or indirectly from interactions between the atmosphere and the underlying surface. The turbulent nature of boundary-layer flows is a vital factor in the efficient exchange of momentum, heat and moisture between the earth's surface below and the 'free' atmosphere above. In general, until fairly recently, designers and users of global atmospheric general circulation models (AGCMs) and operational numerical weather prediction models (NWPMs) have not been concerned with the details of boundary-layer and surface properties and processes in their own right but mainly with the influence they exert on weather systems and circulation characteristics on the much larger, synoptic or even global scales. However, the recent upsurge in the simultaneous developments of three-dimensional AGCMs for the study of climate and climate change, and of increasingly sophisticated and more highly resolved operational NWPMs, has resulted in more effort now being directed towards delineating details in boundary-layer structure and in

* Abridged version of lecture notes prepared for the Centre National d'Études Spatiales (CNES) Summer School of Space Physics, 'Climatology and Space Observations', Roscoff, France, July 1986.

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determining the characteristics of surface climatologies. Studies with AGCMs have indicated considerable sensitivity of their simulations to changes in surface properties such as albedo, soil moisture and surface roughness. Also, some NWPMs now in operational service are expected to forecast the near-surface meteorological variables, and even changes in surface properties. The importance then of 'land-surface processes' and the need to understand and represent them better in AGCMs and NWPMs are now well established.

Following the Joint Scientific Committee Scientific Steering Group on Land-Surface Processes of the World Climate Research Programme (World Climate Programme 1985), I shall adopt the pragmatical definition of land-surface processes as those phenomena which control the fluxes of heat, moisture and momentum between the surface and the atmosphere over the continents. These processes influence both the circulation of the atmosphere, often remotely, and the climate of the surface.

Many important dynamical and physical processes are governed by spatial (and temporal) scales very much smaller than the typical limits of resolution of either a numerical model or an observing system. Such subgrid-scale processes cannot be dealt with explicitly in the models; however, their statistical effects at the resolved scales must be included and are determined in terms of the explicitly resolved variables. This technique is called parametrization and usually introduces empirical terms (parameters) into a model's prescription of the processes. For a fuller discussion of parametrization in numerical models see, for example, Smagorinsky (1982).

My aim here is to introduce the range of subgrid-scale land-surface processes which, it is generally recognized, need to be represented by parametrizations in climate and numerical weather prediction models. Discussion is restricted in the main to non-vegetated land surfaces and focuses in particular on the surface-energy and mass (moisture) fluxes. A more general and fairly comprehensive review of the then current practices in AGCMs was provided by Carson (1982), with an update for Meteorological Office models only in Carson (1986a). As implied above, the parametrization of land-surface processes is a very active field of research and model development, and methods labelled 'current' may quickly become susperseded. New approaches ('schemes') are being developed and tested continuously. A single paper cannot do justice to the range and complexity of tried schemes and unresolved problems even in the apparently restricted topic of land-surface processes. The special characteristics and problems of vegetated land surfaces, ice-covered surfaces and ocean surfaces are not dealt with here. It should be assumed throughout that discussions refer only to non-vegetated, snow-free land surfaces, unless explicitly stated otherwise. Some of the particular problems associated with snow-covered, non-vegetated land surfaces will be described briefly in Part II (Carson 1987).

It should also be stressed that there are many factors in a typical AGCM or NWPM which will have a direct or indirect bearing on the character and performance of the land-surface processes but which are not themselves governed directly by, nor specified explicitly in terms of, surface properties. Obvious examples amongst the other physical parametrizations include components of the radiation scheme, the cloud scheme, the representation of rainfall and snowfall, the delineation of the atmospheric boundary layer and the parametrization of turbulent mixing within it away from the surface, deep convection, etc. A numerical model's general structure with respect to, for example, horizontal domain, spatial and temporal resolutions, distribution and number of surface types, specification of orography, etc. will also determine to some extent the quality of its simulations or predictions of the surface and near-surface climatologies. Such considerations of the general problem of representing the effects of land-surface processes in AGCMs and NWPMs are beyond the scope of this introductory paper.

For convenience, this paper has been divided into two parts. Part I will deal with the general boundary conditions for momentum transfer and the balance equations for energy and mass (moisture) transfer at a bare-soil surface. Also the surface radiative properties and fluxes, and the physical character and parametrization of the surface turbulent exchanges, are considered. Part II (Carson 1987) will

concentrate upon soil heat conduction and the land-surface temperature, and surface hydrology and the soil water budget. A brief description of some of the special features of snow-covered surfaces will also be given.

2. The boundary conditions for momentum, energy and mass (moisture) transfer at a bare-soil surface

A natural and instructive way to delineate and introduce the various land-surface processes of interest is through the boundary conditions for momentum transfer and the balance equations for the energy and mass (moisture) transfer that apply at the surface. Most of the current generation of AGCMs and NWPMs involve such boundary conditions, but with varying degrees of complexity and sophistication in their use, and in the parametrizations chosen to represent individual components of the system. For the moment consider in turn the boundary constraints on and relations between the momentum, energy and mass fluxes.

2.1 *Surface momentum flux*

In an aerodynamic sense the atmospheric boundary layer is simply the lowest layer of the atmosphere under the direct influence of the underlying surface from which momentum is extracted and transferred downwards to overcome surface friction. Thus the aerodynamically rough land surface provides a sink for atmospheric momentum, the removal of which is represented by the viscous drag (i.e. horizontal shearing stress), $\underline{\tau}$, which, by convention, is a vectorial measure of the downward flux of horizontal momentum.

The surface boundary conditions for momentum transfer are the no-slip condition (i.e. the mean horizontal wind is zero at the surface) and the constraint that $\underline{\tau}$ is parallel to the limiting wind as the surface is approached.

2.2 *Surface energy flux balance*

The energy flux balance at a bare-soil surface may be expressed as

$$G_0 = R_N - H - Q \qquad \ldots \; \ldots \; \ldots \; \ldots \; \ldots \; \ldots \quad (1)$$

where R_N is the net radiative flux, H is the turbulent sensible-heat flux, Q is the latent-heat flux due to surface evaporation and G_0 is the flux of heat into the soil. All these terms (units W m^{-2}) are positive when the fluxes are in the directions indicated by the arrows in Fig. 1.

2.3 *Mass flux at the surface*

For our purposes the mass flux at the surface will be taken to be simply the moisture flux expressed as

$$M_0 = P_r - E - Y_0 \qquad \ldots \; \ldots \; \ldots \; \ldots \; \ldots \; \ldots \quad (2)$$

where P_r is the intensity of surface rainfall, E is the surface evaporation rate (turbulent flux of water vapour), Y_0 denotes intensity of surface run-off and M_0 represents the net mass flux of water into the soil layer. As defined, the flux terms in equation (2) strictly have SI units of kg m^{-2} s^{-1}, but it is common to refer to the rates involved in terms of a representative depth (of water) per unit time. The fluxes are positive when they have the directions indicated in Fig. 2.

Since $Q = L_e E$ where L_e is the latent heat of evaporation, the evaporative flux appears explicitly in equations (1) and (2), and thus provides a direct and important coupling between the heat and moisture budgets at the surface.

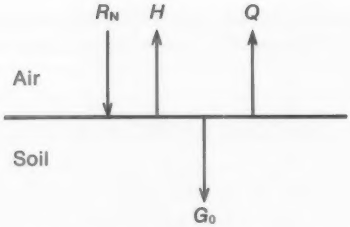

Figure 1. Schematic representation of the energy flux balance at a bare-soil surface (see text for definitions of symbols).

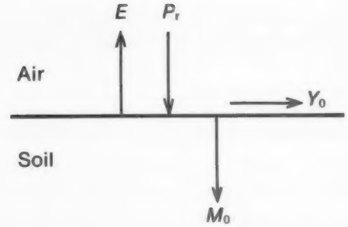

Figure 2. Schematic representation of the mass (moisture) flux balance at a bare-soil surface (see text for definitions of symbols).

2.4 *The parametrization problem*

A knowledge of heat conduction and water transport in the soil is needed to parametrize the terms G_0 and M_0 respectively. In AGCMs and NWPMs this usually leads to the reformulation of equation (1) as a prognostic equation for the surface temperature, and of equation (2) as a prognostic equation for the mass of water stored in a specified depth of surface-soil layer, i.e. the soil moisture content.

The boundary conditions and surface balance equations described above involve a wide range of subgrid-scale physical and dynamical processes in both the atmosphere and the soil. Therefore it is convenient to consider the nature and parametrization of the various individual components separately:

(a) surface radiative properties and fluxes,
(b) surface turbulent exchanges,
(c) soil heat conduction and the land-surface temperature, and
(d) surface hydrology and the soil water budget.

3. Surface radiative properties and fluxes

Since solar radiation provides most of the energy needed to maintain the general circulation of the atmosphere and since the major input of this energy to the earth–atmosphere system occurs at the surface, it seems natural to start a discussion of land-surface processes by considering the surface radiative properties and fluxes. The term R_N in equation (1) acknowledges the importance of, and the need to determine, the net imbalance of radiative fluxes to and from the land surface, expressed here simply as the sum of the net short-wave radiative flux, R_{SN}, and the net long-wave radiative flux, R_{LN}. Therefore, R_N is given by

$$R_N = R_{SN} + R_{LN} . \qquad \ldots \ldots \ldots \ldots \ldots \ldots \quad (3)$$

The components of R_{SN} and R_{LN} are shown schematically in Fig. 3.

3.1 *Surface short-wave radiation balance*

Short-wave radiation is reflected at the earth's surface. Therefore if $R_{S\downarrow}$ is the downward short-wave radiative flux (including both the direct solar flux and diffuse radiation from the sky), the upward short-wave radiative flux is given by $R_{S\uparrow} = \alpha R_{S\downarrow}$ where α is the surface short-wave reflectivity, which is usually called the albedo. The net short-wave radiation, R_{SN}, is the difference between the downward and upward fluxes.

$$R_{SN} = R_{S\downarrow} - R_{S\uparrow} = (1 - \alpha) R_{S\downarrow} . \qquad \ldots \ldots \ldots \ldots \ldots \quad (4)$$

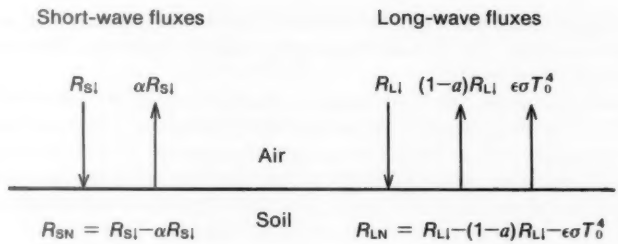

Figure 3. Schematic representation of the short-wave and long-wave radiative fluxes at a bare-soil surface (see text for definitions of symbols).

The albedo depends on the solar zenith angle, the spectral distribution of solar radiation incident on the surface and whether the radiation is direct or diffuse, as well as on the character of the surface as determined by the vegetation (type, density and state), soil type, soil moisture, and whether the surface is snow- or ice-covered. Although generally a long way removed from representing the full complexity of its functional dependence on all such quantities, nevertheless α in AGCMs and NWPMs is usually accorded some variation with the broad character of the surface. In AGCMs it has a specific geographical dependence (see, for example, Carson 1982) and in many models it is still the only land-surface or soil parameter which is given such a geographical variation (see, for example, Carson 1986a).

A good illustration of the current status of the global specification of α suitable for use in large-scale atmospheric models is the recent work of Wilson and Henderson-Sellers (1985) on which is based the distribution of grid-box, snow-free, land-surface albedos used in the Meteorological Office weather forecasting and climate models (see, for example, Carson 1986a). The values range from about 0.12, typical of, for example, the northernmost fringes of land and also tropical forests, to 0.35 used for the most arid, light-coloured desert regions; 0.18 is thought appropriate to the United Kingdom.

3.2 *Surface long-wave radiation balance*

If $R_{L\downarrow}$ is the downward long-wave radiation and a is the surface absorptivity to long-wave radiation, then the net incoming flux from the atmosphere is $aR_{L\downarrow}$. From Stefan's law, the upward flux due to thermal emission at the earth's surface is $\epsilon\sigma T_0^4$, where T_0 is the surface temperature, ϵ is the long-wave emissivity at the surface and σ is the Stefan–Boltzmann constant. Therefore the net long-wave radiative flux, R_{LN}, is given by

$$R_{LN} = aR_{L\downarrow} - \epsilon\sigma T_0^4 . \qquad \ldots \ldots \ldots \ldots \ldots \ldots \quad (5)$$

It is common practice to simplify equation (5) by combining the definition of ϵ with Kirchhoff's law to give $a = \epsilon$. Equation (5) then reduces to

$$R_{LN} = \epsilon(R_{L\downarrow} - \sigma T_0^4) . \qquad \ldots \ldots \ldots \ldots \ldots \ldots \quad (6)$$

It is known that ϵ has a wavelength dependence and that it varies according to the character of the surface as discussed, for example, by Buettner and Kern (1965), Kondratyev (1972), Paltridge and Platt (1976) and Kondratyev *et al.* (1982). Values quoted for ϵ range from 0.997 for wet snow to 0.71 for quartz. Kondratyev *et al.* (1982) comment that, on average, ϵ for natural surfaces lies within the range

0.90–0.99 and they cite several authors who have inferred that 0.95 may be assumed as the mean relative emissivity of the earth's surface. They do caution, however, that the problem of measuring the emissivity of natural surfaces is far from solved.

Although there are exceptions, the most common practice in AGCMs and NWPMs is to assume explicitly or implicitly that all surfaces act like perfect black bodies for long-wave radiation with $\epsilon = 1$. To a large extent this simply reflects the preoccupation of numerical modellers with other apparently more important and immediate problems with their physical parametrizations. It is likely that the increasing complexity and sophistication of land-surface descriptions in models will also generate more critical and discriminatory approaches to the specification of ϵ.

3.3 *Evaluation of R_N*

The net radiation, R_N, is found by substituting equations (4) and (6) into (3) to give

$$R_N = (1-\alpha)R_{S\downarrow} + \epsilon(R_{L\downarrow} - \sigma T_0^4)\,. \qquad \text{(7)}$$

The parametrization of $R_{S\downarrow}$ and $R_{L\downarrow}$ is beyond the scope of this paper; they are not formally classed as land-surface processes and may be regarded here as externally given forcing factors. It should be stressed though that a correct evaluation of $R_{S\downarrow}$ and $R_{L\downarrow}$ is a crucial element in establishing sensible energy and moisture balances at the surface.

From equation (7) it is clear that once T_0 and the downward fluxes of short-wave and long-wave radiation are known, the net radiation can be determined if ϵ and α are specified.

4. Surface turbulent exchanges

4.1 *Definition of the surface turbulent fluxes*

The atmospheric boundary layer (sometimes called the planetary boundary layer or mixing layer) is the lowest layer of the atmosphere under the direct influence of the underlying surface. The flow in this layer is turbulent except possibly in very stable conditions (e.g. at night in the presence of strong surface-based temperature inversions). The velocity, temperature, humidity and other properties in a turbulent flow can be considered as random functions in space and time, and it is usually necessary to resort to a statistical approach for the calculation of many boundary-layer properties. In particular this introduces the concepts of mean values, fluctuations and variances into the description of the turbulent properties of the flow. For example, if ξ is some conservative quantity which fluctuates because of the turbulent motion, then it is usually written as

$$\xi = \overline{\xi} + \xi' \qquad \text{(8)}$$

where $\overline{\xi}$ is some suitably defined mean value of ξ, and ξ' is called the turbulent or eddy fluctuation (see schematic illustration in Fig. 4).

In the notation of equation (8), the term $\overline{w'\xi'}$ represents the eddy covariance of ξ and the vertical component of the flow, w, and denotes the vertical turbulent flux of ξ at a given height in the atmospheric boundary layer. Let

$$F_\xi = (\overline{w'\xi'})_0 \qquad \text{(9)}$$

denote the surface value of this flux.

In the context of this discussion about land-surface processes, it is the turbulent fluxes of momentum, $\underline{\tau}$, sensible heat, H, and water vapour, E, that are of particular interest and are given by

$$\underline{\tau} = \rho\{-(\overline{w'u'})_0, -(\overline{w'v'})_0\} \quad \text{... ... ... (10)}$$

$$H = \rho c_p(\overline{w'\theta'})_0 \quad \text{... ... ... (11)}$$

$$E = \rho(\overline{w'q'})_0 \quad \text{... ... ... (12)}$$

where u and v are the components of the horizontal wind, θ is the potential temperature, q is the specific humidity, ρ is a representative mean air density near the surface and c_p is the specific heat of air at constant pressure. The direction of $\underline{\tau}$ is determined by the limiting wind direction as the surface is approached.

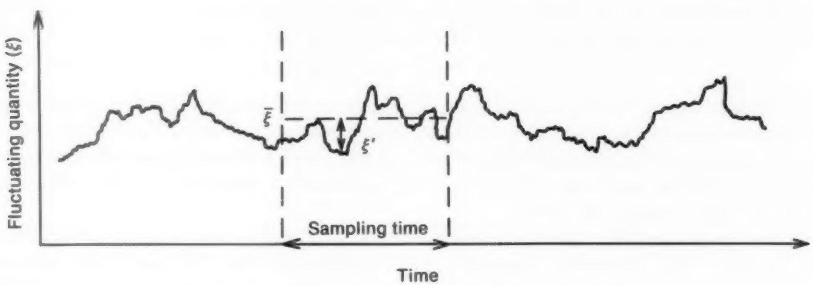

Figure 4. Schematic representation of the mean value, $\bar{\xi}$, and the eddy fluctuation, ξ', determined for a particular sampling time from a time trace of the fluctuating quantity, ξ.

4.2 *The bulk-aerodynamic formulae*

It is standard practice, particularly in AGCMs and NWPMs, to represent the mean vertical turbulent flux, F_ξ, by

$$F_\xi = -C_\xi V(z_l)\Delta\xi(z_l) \quad \text{... ... ... (13)}$$

where $\Delta\xi(z_l) = \xi(z_l) - \xi_0$, z_l is some specified height above the surface but within the boundary layer, $\xi(z_l)$ and $V(z_l)$ are the mean values of ξ and the horizontal wind speed at z_l, and ξ_0 is the surface value of ξ. (Note that the bar notation to denote mean values has been omitted to simplify the symbolism.) The bulk transfer coefficient, C_ξ, defined in a strictly mathematical sense by equation (13), is a complicated function of height, atmospheric stability, surface roughness and, for a vegetated surface, of other physical and physiological characteristics of the vegetation.

Without loss of generality, z_l may be taken as the notional height of a particular numerical model's first level above the underlying surface.

In bulk-aerodynamic form, the surface turbulent fluxes of equations (10) to (12) are

$$\underline{\tau} = \rho C_D V(z_l)\underline{V}(z_l) \quad \text{... ... ... (14)}$$

$$H = -\rho c_p C_H V(z_l)\{\theta(z_l) - \theta_0\} \quad \text{... ... ... (15)}$$

$$E = -\rho C_E V(z_l)\{q(z_l) - q_0\} \quad \text{... ... ... (16)}$$

where C_D, C_H and C_E are the bulk transfer coefficients for momentum transfer, heat transfer and water

vapour transfer (note that C_D is the traditional drag coefficient), and θ_0 and q_0 are the surface values of the potential temperature and specific humidity.

To determine the fluxes from equations (14) to (16) C_ξ must be prescribed or expressed in terms of modelled variables and parameters and, in addition to the variables modelled explicitly at z_l, θ_0 (simply related to the surface temperature T_0) and q_0 need to be known. The prediction of T_0 is usually based on knowledge of heat transfer through the soil, but the prediction of q_0 is not so easy since its implied value is inextricably linked to the parametrization of surface hydrology.

A related approach to equation (13) for the representation of the turbulent fluxes at natural surfaces is the so-called resistance approach. Turbulent transfer in the atmospheric boundary layer is seen as a process analogous to the flow of electric current and, in the spirit of Ohm's law, F_ξ is written as

$$F_\xi = -\frac{\Delta\xi}{r_\xi} \qquad \dots\dots\dots\dots\dots\dots (17)$$

where, in a similar manner to C_ξ in equation (13), equation (17) can be regarded as defining r_ξ, the aerodynamic resistance to the 'flow' of F_ξ. Comparison of equations (13) and (17) yields $r_\xi = \{C_\xi V(z_l)\}^{-1}$. The resistance approach has particular appeal when dealing with the complicated and multiple routes for sensible heat transfer and evaporation from vegetated surface (see, for example, Monteith 1965, Perrier 1982 and Rosenberg *et al.* 1983).

For a discussion of the large variety of specifications of C_ξ then in current use in AGCMs see, for example, Carson (1982). However, discussion here is limited to the approach most acceptable to boundary-layer experts and increasingly more prevalent in the current generation of AGCMs and NWPMs — the use of the Monin–Obukhov similarity theory.

4.3 *Use of the Monin–Obukhov theory of the surface-flux layer to determine the bulk transfer coefficients*

Adjacent to the surface is a shallow layer in which the turning of the wind with height may be ignored and the vertical fluxes of momentum, heat, and water vapour may be approximated closely by their surface values (i.e. for many practical purposes the turbulent fluxes in this layer may be assumed to be virtually constant with height). This layer is often referred to as the constant-flux layer, but this terminology can mislead the unwary (e.g. the fluxes generally have their largest vertical gradients at the surface) and so it is better to use the more appropriate term of surface-flux layer.

The Monin–Obukhov similarity hypothesis for the surface-flux layer is the most widely accepted approach for describing the properties of the surface layer. Brought down to the very simplest terms, similarity methods depend on the possibility of being able to express the unknown variables in non-dimensional form, there being suitable arguments for saying that there exist a length scale, velocity scale (or time-scale) and temperature (and humidity) scale relevant in doing this. The non-dimensional forms are then postulated to be universal in character and this will hold for as long as the scales remain the relevant ones.

From the Monin–Obukhov theory of the surface-flux layer (see Appendix for details) it can be shown that

$$C_\xi = C_\xi(z_l/z_0, Ri_B)$$

where z_0 is the surface roughness length and Ri_B is a bulk Richardson number for the surface layer.

Over bare soil z_0 is a characteristic of the surface and is usually independent of the flow. There are also corresponding characteristic 'surface roughness lengths' for heat and water vapour transfer. The

problems of evaluating effective areal roughness lengths and of discriminating between them for the different properties are complex, and it remains common practice to use the same estimate of z_0 when computing all three bulk transfer coefficients.

The bulk Richardson number, Ri_B, is a measure of the stability of the surface layer. In stable conditions $Ri_B > 0$, whereas $Ri_B < 0$ in unstable conditions; $Ri_B = 0$ when the surface layer is neutral. The Ri_B can be computed easily from the model variables using

$$Ri_B = \frac{gz_l}{T} \frac{\{\Delta\theta(z_l) + 0.61\, T\Delta q(z_l)\}}{V^2(z_l)}$$

where g is the acceleration due to gravity and T is a representative air temperature in the surface layer. To illustrate the behaviour of bulk transfer coefficients derived from the Monin–Obukhov theory, Fig. 5 shows how C_D, C_H and C_E used in the Meteorological Office 11-layer AGCM vary with Ri_B. In that particular model $z_l = 100$ m, over land $z_0 = 0.1$ m and over the sea $z_0 = 10^{-4}$ m.

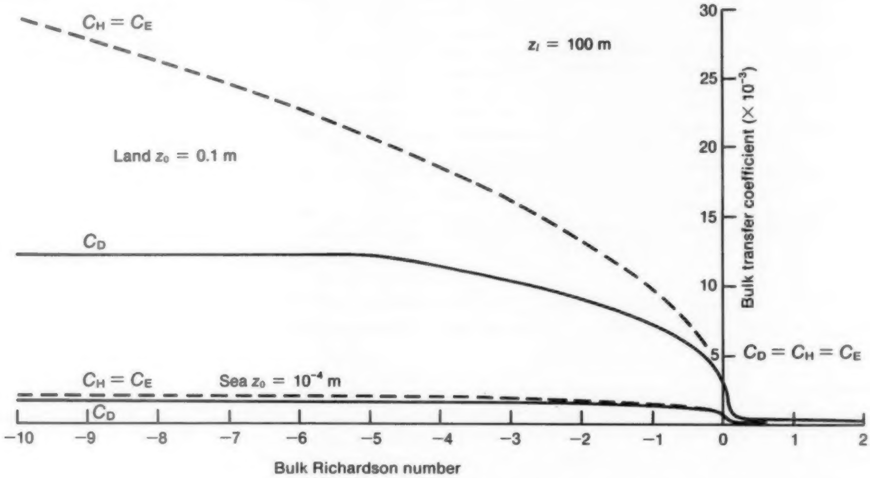

Figure 5. Surface-layer bulk transfer coefficients derived from the Monin–Obukhov similarity theory and used in the Meteorological Office 11-layer AGCM.

4.4 *Surface roughness length*

The surface roughness length, like the surface albedo, is a land-surface characteristic which has a marked geographical variation. In most of the current generation of AGCMs and NWPMs, z_0 has direct and indirect effects on the surface turbulent exchanges of sensible heat and moisture, as well as on the surface shearing stress. However, the evaluation of an effective areal surface roughness length for heterogeneous terrain is an important practical issue that poses a variety of, as yet unsatisfactorily resolved, problems.

The effective areal z_0 for natural surfaces is rarely estimated from the wind profile and/or surface-shear stress measurements. Instead, it is most likely to be determined indirectly from a knowledge of, for example, terrain relief (elevation, slope, etc.), land use, and type and distribution of the surface roughness elements. Algorithms, however qualitative, are needed to perform this function sensibly, at

least in a fairly local (1 km × 1 km) sense. The pros and cons of alternative approaches to the question of how to average over larger areas have been discussed by Carson (1986b).

Most standard boundary-layer textbooks provide a table of values of z_0 as a function of terrain type described qualitatively in terms of relief and vegetation characteristics. Such traditional relationships may well be adequate on the very local scale for the smoother, quasi-homogeneous types of terrain, but can be expected to be less well founded for areal averages over rough, heterogeneous terrain typical of, say, a European semi-rural landscape with small hills, woods, fields, crops, hedges, towns, lakes, etc. Wieringa (1986) has addressed this problem and produced a table giving effective areal z_0 in terms of a terrain classification when there are no significant orographic features (see Table I).

Table I. *Effective mesoscale surface roughness length, z_o, expressed as a function of land use and proposed by Wieringa (1986); h is the height of the major surface obstacles.*

Land use category	z_0 *metres*
Sea (minimal fetch 5 km)	0.0002
Small lake, mud flats	0.006
Morass	0.03
Pasture	0.07
Dunes, heath	0.10
Agriculture	0.17
Road, canal (in Dutch landscape, tree lined)	0.24
Orchards, bushland	0.35
Forest	0.75
Residential built-up area ($h \leqslant 10$ m)	1.12
City centre (high-rise building)	1.6

For a fuller discussion of issues concerning the evaluation of effective z_0 see recent papers by Smith and Carson (1977), Mason (1986), Carson (1986b), Wieringa (1986) and André and Blondin (1986).

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Appendix — Use of the Monin–Obukhov similarity theory

A.1 *The Monin–Obukhov similarity theory*

The Monin–Obukhov similarity hypothesis for the fully turbulent surface-flux layer (where the Coriolis force is neglected) states that for any transferable property, the distribution of which is homogeneous in space and stationary in time, the vertical flux–profile relation is determined by the parameters

$$\frac{g}{T}, \ \frac{|\underline{\tau}|}{\rho}, \ \frac{H}{\rho c_p}, \ \frac{E}{\rho}$$

where g/T is the Archimedean buoyancy parameter, g is the acceleration due to gravity and T is a representative air temperature in the surface layer.

It is convenient to introduce scaling parameters u_*, θ_* and q_* which are defined in terms of $\underline{\tau}$, H and E by $|\underline{\tau}| = \rho u_*^2$, $H = -\rho c_p u_* \theta_*$ and $E = -\rho u_* q_*$ (in general, the surface turbulent flux of ξ is given by

$F_\xi = -u_*\xi_*$). Using these three equations means that the above set of four parameters is equivalent to the set

$$\frac{g}{T},\ u_*,\ \theta_*,\ q_*$$

where θ_* and q_* can be combined to give

$$\psi_* = \theta_* + 0.61\,Tq_*$$

which is akin to a virtual potential temperature scaling value.

Instead of using the buoyancy parameter g/T it is convenient to use the length scale L, called the Monin–Obukhov length, defined uniquely by g/T, q_* and ψ_*.

$$L = \frac{Tu_*^2}{kg\psi_*} = \frac{-\rho c_p T u_*^3}{kg(H + 0.61 c_p TE)}\ .$$

By convention Kármán's constant ($k \approx 0.4$) is introduced solely as a matter of convenience. In the surface-flux layer, L is effectively constant. The turbulent flow is classed as unstable when $L < 0$ (i.e. when the net surface-buoyancy flux is positive), stable when $L > 0$ (i.e. when the net surface-buoyancy flux is negative), and neutral when $|L| \rightarrow \infty$ (i.e. when the net surface-buoyancy flux is zero).

Now L, u_*, θ_* and q_* may be taken as the set of basic parameters which uniquely determine the relationships between the surface-layer vertical gradients of wind, potential temperature and specific humidity to the corresponding surface turbulent fluxes. Dimensional analysis leads to the vertical flux–gradient relationship expressed in the general form

$$\frac{\partial \xi}{\partial z} = \frac{\xi_*}{kz}\,\phi_\xi(z/L) \qquad \text{...(A1)}$$

where z is height above the surface. It is then hypothesized that $\phi_\xi(z/L)$ is a universal function of z/L only, which may be of different form for each mean transferable property, ξ. The form of the functions has to be determined empirically from analysis of surface-layer data; the overall observational evidence is that they decrease with unstable stratification ($L < 0$) and increase with stable stratification ($L > 0$).

For specified functions for ϕ_ξ, equation (A1) can be integrated to provide flux–profile relationships for the surface layer:

$$\frac{k\{\xi(z) - \xi(z_r)\}}{\xi_*} = \int_{\zeta_r}^{\zeta} \frac{\phi_\xi(\eta)}{\eta}\,\mathrm{d}\eta = \Phi_\xi(\zeta, \zeta_r). \qquad \text{...(A2)}$$

where $\zeta = z/L$ and $\zeta_r = z_r/L$, and z_r is some reference height at which ξ is known. In practice, equation (A2) used in conjunction with the basic equations relating surface fluxes to the scaling parameters can be used to estimate the surface turbulent fluxes of momentum, heat and moisture from a knowledge of the corresponding surface-layer profiles of wind, potential temperature and humidity.

A.2 *The similarity functions*

The general character of the similarity functions is fairly well established over a limited range of stability conditions centred on neutral, but their specification for extreme stability conditions (both stable and unstable) is much more debatable and uncertain. The general behaviour of the functions is

that ϕ_ξ increases with increasing stability — decreasing turbulence decreases the mixing and hence increases the normalized gradient of ξ. Fig. A1 illustrates schematically the changing character of the surface-layer wind profile throughout a clear day and clear night. For details see, for example, chapter 6 of Panofsky and Dutton (1984).

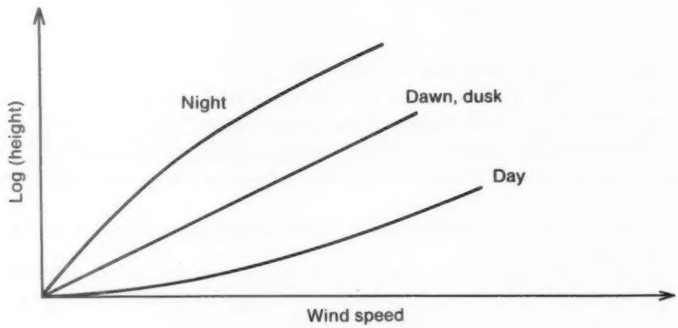

Figure A1. Schematic representation of the diurnal variation of the surface-layer wind profile.

There is a variety of postulated, empirical forms of ϕ_ξ (see, for example, chapter 6 of McBean *et al.* 1979). The following examples have been selected subjectively, but are typical of the type of formulae commonly adopted as the basis of parametrizations for the turbulent fluxes in numerical models.

(a) *Unstable and neutral conditions ($z/L \leqslant 0$).* The Dyer and Hicks (1970) formulae are

$$\phi_H = \phi_E = \phi_M^2 = (1-16z/L)^{-\frac{1}{2}} \qquad 0 \geqslant z/L \geqslant -1$$

where ϕ_M, ϕ_H and ϕ_E are the respective ϕ_ξ for the turbulent transfers of momentum, sensible heat and water vapour. Since these formulae are limited to when $|z/L| \leqslant 1$, other empirical approaches may need to be invoked for more unstable conditions. For a particular choice of extrapolation beyond the Dyer and Hicks limit towards the free-convection limit see Carson (1982, 1986a).

(b) *Stable conditions ($z/L > 0$).* The formulae proposed by Webb (1970) are

$$\phi_H = \phi_E = \phi_M = \begin{cases} 1 + 5z/L & 0 < z/L \leqslant 1 \\ 6 & 1 < z/L < 6 \end{cases}$$

The problem of extending the functional form of ϕ_ξ to highly stable conditions was discussed by Carson and Richards (1978).

A.3 C_ξ *from the Monin–Obukhov similarity theory*

The bulk transfer coefficients introduced in section 4.2 can be derived from the Monin–Obukhov similarity theory. The surface flux, F_ξ, can be written in terms of u_* and ξ_*, and also in terms of C_ξ, namely

$$F_\xi = -u_*\xi_* = -C_\xi V(z_l)\Delta\xi(z_l)$$

which yields

$$C_\xi = \left(\frac{u_*}{V(z_l)}\right)\left(\frac{\xi_*}{\Delta\xi(z_l)}\right). \qquad \text{... ... ...(A3)}$$

For the Monin–Obukhov theory to be appropriate z_l must be fully within the surface layer so that equation (A2) can be invoked in the particular form

$$\frac{k\Delta\xi(z_l)}{\xi_*} = \int_{\zeta_\xi}^{\zeta_l} \frac{\phi_\xi(\eta)\,\mathrm{d}\eta}{\eta} = \Phi_\xi(\zeta_l, \zeta_\xi) \qquad \ldots \ldots \ldots \ldots \ldots \text{(A4)}$$

where $\zeta_l = z_l/L$, and $\zeta_\xi = z_\xi/L$ is defined such that $\xi(z_\xi) = \xi_0$.

The nature of the similarity formulation implies a logarithmic singularity in Φ_ξ as $z \rightarrow 0$. This is avoided by defining the level z_ξ as the virtual height at which the ξ-profile, defined by equation (A2) and extrapolated towards the surface, attains the actual surface value ξ_0. For momentum transfer this level, denoted by z_0 and called the surface roughness length, is defined as the virtual height at which $V = 0$ on the postulated wind profile.

From equations (A3) and (A4), C_ξ can be specified in terms of finite integrals of the Monin–Obukhov similarity functions, thus

$$C_\xi = k^2\, \Phi_M^{-1}(\zeta_l, \zeta_0)\, \Phi_\xi^{-1}(\zeta_l, \zeta_\xi) \qquad \ldots \ldots \ldots \ldots \ldots \text{(A5)}$$

where

$$\Phi_M(\zeta_l, \zeta_0) = \int_{\zeta_0}^{\zeta} \frac{\phi_M(\eta)\,\mathrm{d}\eta}{\eta} = \frac{kV(z)}{u_*}$$

and $\zeta_0 = z_0/L$. In general, with ϕ_ξ specified as discussed earlier, equation (A5) gives C_ξ as a function of ζ_l, ζ_0 and ζ_ξ.

It is generally more convenient for modelling purposes to express C_ξ directly as a function of the explicitly modelled variables $V(z_l)$ and $\Delta\xi(z_l)$. This can be achieved by using a bulk Richardson number for the surface layer, Ri_B, instead of ζ_l as the stability indicator. In terms of model variables Ri_B is given by

$$Ri_B = \frac{gz_l}{T} \frac{\{\Delta\theta(z_l) + 0.61\,T\Delta q(z_l)\}}{V^2(z_l)}$$

which can be related to ζ_l through

$$Ri_B = \frac{\zeta_l C_D^{3/2}}{kC_H}$$

For a full description of the method and assumptions made see, for example, Carson and Richards (1978).

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A comparison of numbers of visually estimated and instrumentally measured wind observations from merchant ships

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Summary

A comparison is made of the numbers of reports of visually estimated and instrumentally measured winds stored in the Meteorological Office main marine data bank. The geographical distribution of the two types of observation is analysed and the effects are discussed.

1. Introduction

The Meteorological Office has a marine data bank (Shearman 1983) containing 55 million observations collected from 1854 to 1984. This is the main source of information for climatological analyses and therefore it is important to be aware of any limitations or inconsistencies in the contents of the data bank.

Observations of wind speed are used more frequently than those of any other variable. There are two types of wind speed observation in the marine data bank, namely those that are visually estimated and those that are instrumentally measured. Visually estimated wind speeds are derived by assessing the state of sea, a particular Beaufort force corresponding to a particular sea state. Before 1960 the majority of wind speeds reported by ships were visually estimated, but in recent years anemometers of various types have become more widely used and the bank now contains a mixture of the two types of observation. The majority of the more recent reports specify the method of observation by the inclusion of a coded indicator.

Instrumental observations may be made using hand-held or permanently mounted anemometers and give the wind speed in knots or metres per second. Although instrumental observations may be thought to have a greater accuracy than visual estimates, such wind speeds should be treated with caution because there is always the possibility that the anemometer was not properly exposed. The wind speed is nearly always affected by the air flow over the ship, particularly in the case of hand-held anemometers where the observer may be sheltered from a wind from a certain direction by the superstructure of the vessel. The speed of the ship and the height of the wind sensor above sea level are additional complications which may or may not have been taken into consideration before the reports were despatched.

Some Voluntary Observing Fleets (VOFs) appear to be changing to instrumentally measured wind speeds as a matter of policy. It is inevitable that the data bank will contain a mixture of visual and instrumental wind speeds. In this paper the proportions of each type of data are described, together with any underlying national trends, and an assessment is made of the consequences of the data mix when carrying out a standard climatological analysis such as the derivation of extremes.

2. The data base

Marine meteorological observations are available in the Meteorological Office marine data bank from 1854 to 1984; most of the observations have been made by deck officers during the course of their duties aboard merchant vessels of the VOF. Data are also received from ocean weather ships, light-vessels, gas and oil rigs, and buoys.

In the early 1960s nine countries were nominated by the World Meteorological Organization to act as collecting centres for meteorological observations from specified ocean areas of the world and to be responsible for archiving the data from their nominated areas of responsibility. Each of the eight countries actively remaining in the scheme (see Fig. 1) has a complete archive of data for its own area and, to some extent, an incomplete archive for the rest of the world based upon observations from ships of its own VOF.

Additionally the United Kingdom has acquired further data, by exchange or purchase, from each of the other data centres with the exception of the USSR and India, although the data from India will become available when they have been processed. The marine data bank is complete up to the end of 1981 for all observations, except those from these two centres, and for some centres until the end of 1984. However, there is a known shortfall of data for the US area of responsibility from 1965 to 1972 which will be remedied in the near future.

Since 1960 an indicator should have been added to the coded observations to distinguish between instrumental and visual data. However, it is possible that an error occurred in coding the indicator for data from the US VOF because there are apparently no instrumentally measured wind speeds reported from 1975 to 1981, although prior to 1975 and since 1981 a large number were reported.

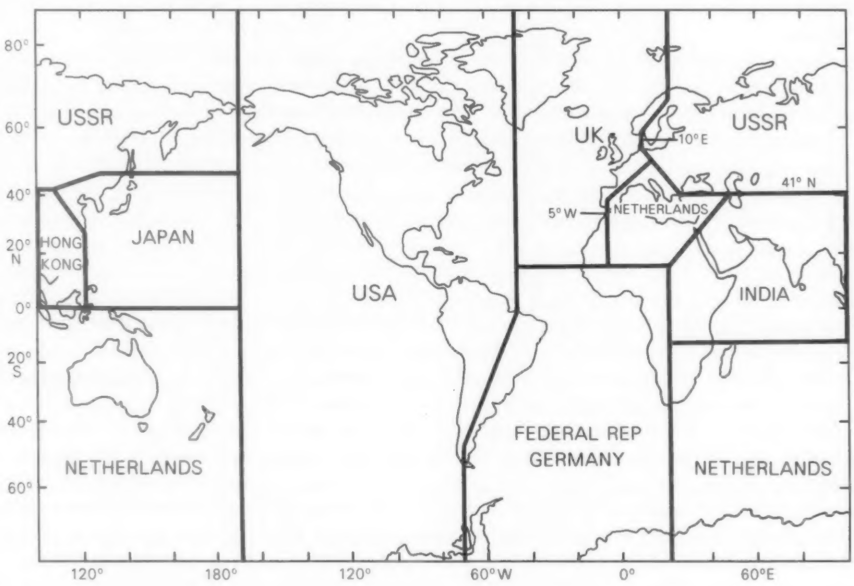

Figure 1. Areas of responsibility of the eight countries acting as collecting centres for meteorological observations.

3. Analysis and results

The two different types of wind data were compared for areas which lay on major shipping routes so that a significantly large sample of observations could be analysed. Reports were examined from vessels recruited by 16 countries which possessed large VOFs, for the period 1854–1984. The numbers of

visually estimated and instrumentally measured wind speeds were calculated for 5-year periods, starting from 1960, for the eight countries of origin shown in Table I.

It can be seen that ships recruited by India have only reported visually estimated wind speeds. The number of measured wind speed reports made little contribution to the overall total of reports for vessels registered in the Federal Republic of Germany, the Netherlands and Yugoslavia. During the late 1960s around a quarter of wind observations from ships of the UK VOF were reported as measured, but since then the total has decreased to between 5 and 10%. Figures for American-recruited vessels indicate that up to a quarter of their observations are measured but since the coding error was corrected in 1982 around a half of the observations received have been measured. The proportion of measured wind speeds reported by France and, in particular, Japan increased markedly during the period since 1960, and Japan now reports 90–100% instrumental wind speeds. The decrease in the total number of observations is due in part to a slight decline in the size of the VOF.

Table I. *Numbers of instrumental (I) and visual (V) wind speeds, by country of origin and year (expressed in units of 100 observations)*

	1960–64		1965–69		1970–74		1975–79		1980–84	
	I	*V*	*I*	*V*	*I*	*V*	*I*	*V*	*I*	*V*
USA	53	859	416	1163	414	1097	—	2500	177	820
France	—	210	56	291	220	176	189	92	98	31
Federal Republic of Germany	1	406	84	780	7	770	22	638	49	1209
India	—	59	—	86	—	68	—	—	—	—
Japan	27	2789	511	1274	993	138	883	31	375	5
The Netherlands	0	530	3	672	—	567	8	419	14	436
United Kingdom	9	430	28	98	97	701	84	1081	118	1030
Yugoslavia	—	31	—	141	—	90	—	41	1	11

To illustrate the geographical effect of individual countries' practices in the use of anemometers, maps have been produced showing the percentage of instrumentally measured winds, covering the area 50° S to 60° N and 180° W to 180° E, for four consecutive 5-year periods between 1960 and 1979, using data from all operating countries. The map for 1960–64 is shown in Fig. 2. Values are plotted at the centre of each 10-degree square; squares containing less than 100 observations have been left blank.

On examining this map it is clear that only 1–2% of the wind speeds reported in the North Atlantic during the period 1960–64 were measured. Similar maps for 1965–69, 1970–74, and 1975–79 are shown in Figs 3, 4 and 5 respectively; it can be seen that in the period 1965–69 the percentage rose to approximately 10% and thereafter remained fairly steady. Data from the Indian Ocean followed a similar pattern, but the percentage rose slightly during the period 1975–79 to about 15%.

In the Mediterranean area approximately 6–7% of reported wind speeds between 1960 and 1964 were measured, rising to about 20% and 35% for the following two periods respectively and decreasing to about 15% in 1975–79.

The number of measured wind speeds reported from the South Atlantic increased steadily from 5% of the total in the period 1960–64 to 20% in the period 1975–79. Along the coast of South America the percentage of measured wind speeds appeared generally higher than in the open seas. This may be due to a ship with an anemometer working continuously in the area.

The Southern Ocean (off the coast of Antarctica) had a consistently higher percentage of measured wind speeds over the 20-year period, although the number of observations was small. This almost

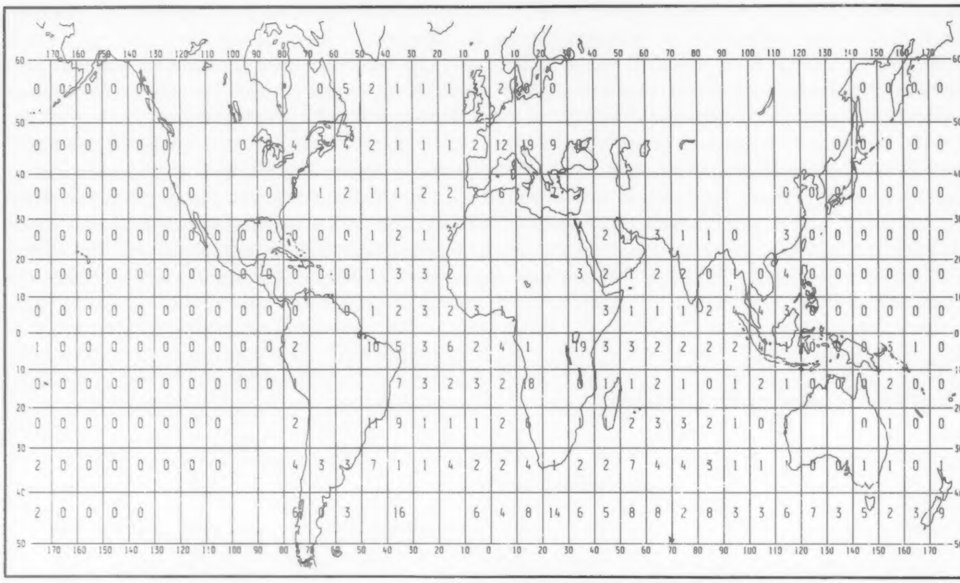

Figure 2. Percentage of instrumentally measured wind observations for the period 1960–64.

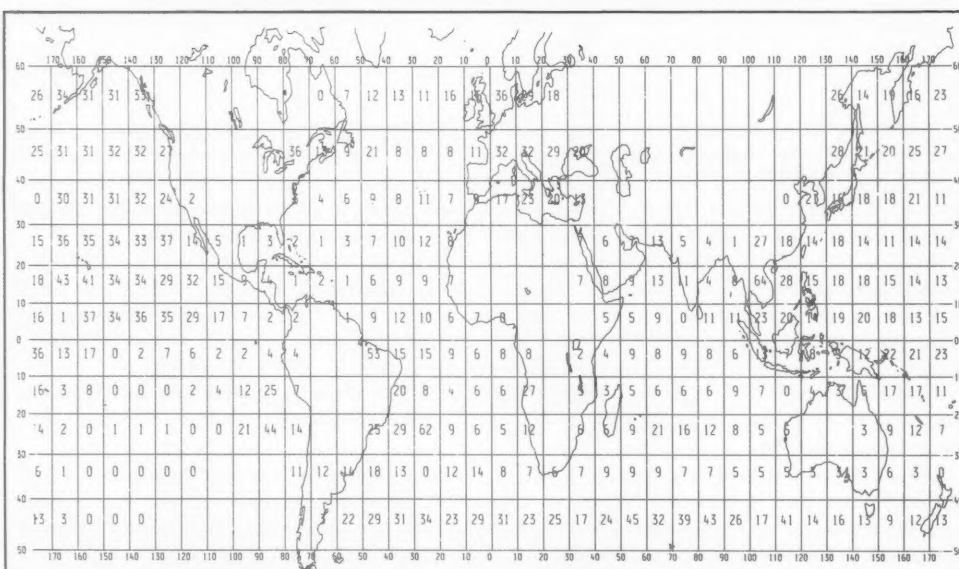

Figure 3. Percentage of instrumentally measured wind observations for the period 1965–69.

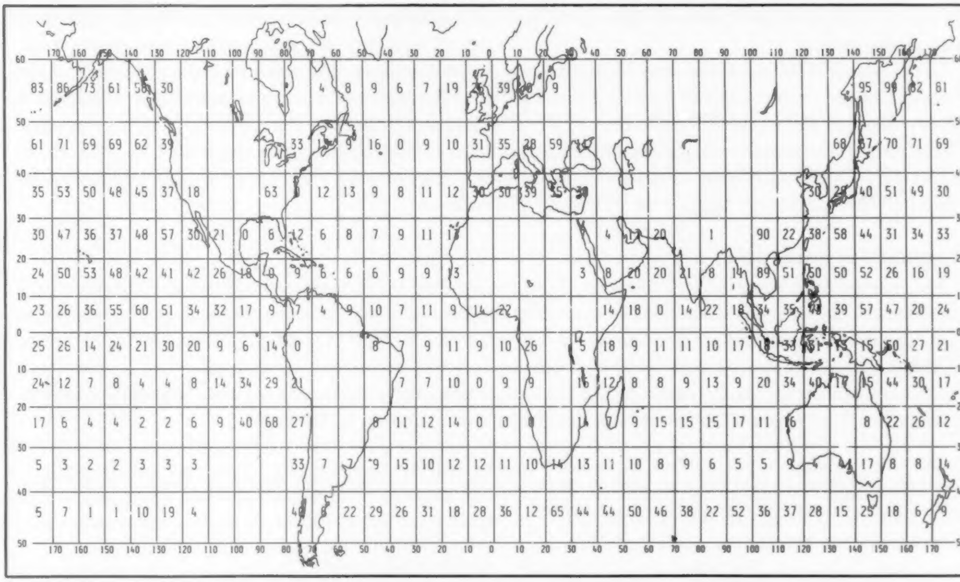

Figure 4. Percentage of instrumentally measured wind observations for the period 1970–74.

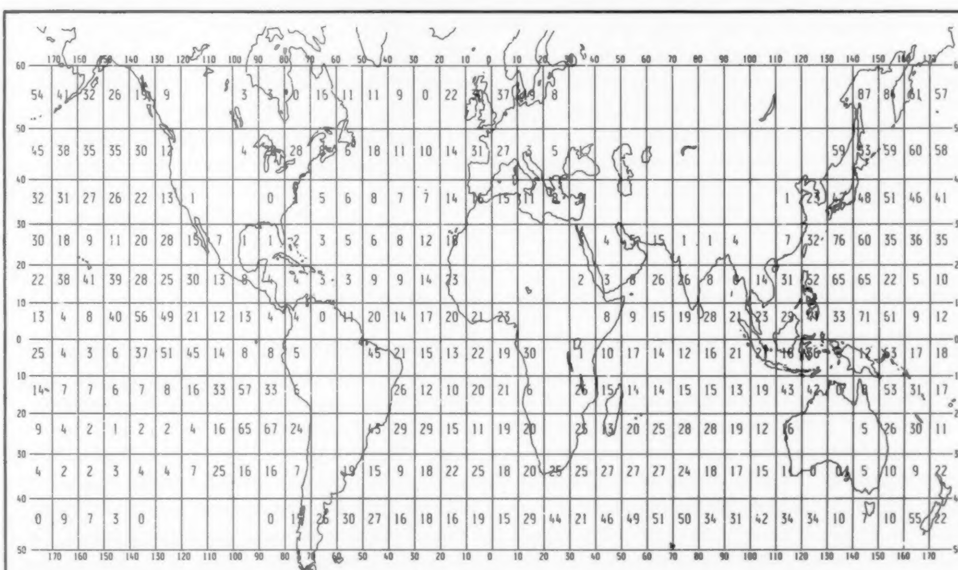

Figure 5. Percentage of instrumentally measured wind observations for the period 1975–79.

certainly reflects the anomalously high contribution from research vessels which are usually equipped with anemometers.

In the period 1960–64 there were virtually no measured wind speeds reported from the whole of the Pacific Ocean. In the second period, from 1965 to 1969, the percentage of measured wind speed observations in the western Pacific rose to 15–20% and in the eastern Pacific to 30–40% of the total. Thereafter, in the eastern Pacific the percentage rose to 40–50% reducing to about 40% in 1975–79. In the western Pacific approximately half the wind speed observations were reported as measured for the period 1970–74, and this figure rose to 60% in 1975–79.

4. Analysis of mixed data

The marine data bank is extensively used to provide data for climatological analyses; therefore it is important to consider the effect on climatological analyses of using a mixture of visual and instrumental data. The area chosen for such an assessment extends from 10 to 20° N and from 130 to 140° E, and since 1970 has contained approximately 50% measured and 50% estimated wind speeds. The frequency of occurrence of instrumentally measured and visually estimated wind speeds for each 'scientific Beaufort scale' point, for the period 1970–74, was calculated and is shown in Table II.

Table II. *Numbers of measured and visual wind speeds, by scientific Beaufort force, for the area 10–20°N, 130–140°E, for the period 1970–74*

Scientific Beaufort force	Number of observations		
	Measured	Visual	Total
0	563	182	745
1	654	622	1 276
2	2 425	1 960	4 385
3	4 596	4 196	8 792
4	5 936	6 292	12 228
5	3 340	4 032	7 372
6	1 474	2 379	3 853
7	619	700	1 319
8	210	192	402
9	35	26	61
10	9	12	21
11	3	3	6
12	1	0	1
Total	19 865	20 596	40 461

For wind speeds of force 3 and below there were more measured than visually estimated wind speeds; for force 4–7 this relationship was reversed; and for force 8 and above the difference was minimal. This confirms that instrumental measurements and visual estimates are providing differing climatologies.

The overall effect of this difference can be best illustrated by deriving extreme values from the two sources of data. This was done by fitting a Weibull function (Weibull 1951) to the distributions of both the measured and visually estimated wind speeds. Extreme values were estimated for return periods, of 1, 2, 5, 10, 20, 50, 100 and 200 years and are presented in Table III. Extremes derived from the measured wind speeds are 3–5 knots lower than those from the visually estimated wind speed distribution. The so-called scientific Beaufort scale (World Meteorological Organization 1970) was used in the calculation. A Weibull analysis of the mixed distribution gave extreme values which were 1–2 knots lower than the values obtained using only visually estimated wind speeds.

Table III. *Extreme values, calculated by fitting a Weibull function to the distributions of the measured and visual wind speeds, using the scientific Beaufort scale*

Return period	Measured wind speed	Visual wind speed	Combined wind speed
years	*knots*	*knots*	*knots*
1	45.4	48.3	47.6
2	46.8	49.9	49.2
5	48.5	52.1	51.1
10	49.8	53.6	52.6
20	51.0	55.1	54.0
50	52.6	57.0	55.8
100	53.7	58.3	57.0
200	54.8	59.7	58.3

5. Conclusion

It appears from the results shown that a few countries are reporting a very large percentage of their wind speed observations as instrumentally measured and it therefore follows that the areas in which the VOF of these countries predominantly operate will also have a higher percentage of measured wind speeds than other areas. For example, the percentages of measured wind speeds reported by Japan and the USA are amongst the highest. Both these countries operate vessels predominantly in the Pacific and the South China Sea and it is these areas which contain the largest percentages of measured wind speeds. Therefore, geographical location should be taken into account when considering the strengths and weaknesses of the data bank in application to marine climatology.

The differences in extreme wind speeds, estimated by fitting a Weibull function to distributions of visual and measured wind speeds, are very significant. They are large enough to lead to inadvertant 'under-designing' if measured wind speeds are used without due care. For example, whereas visual estimates indicate that the return period of a 50-knot wind is about 2 years for an area in the South China Sea, instrumental measurements produce a return period of greater than 10 years. Any adverse effect could also be magnified when wind data are combined with wave or other data to establish the design criteria for structures such as oil rigs, since high waves which add to the structural loading normally occur in association with high winds.

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The Meteorological Office Historical Sea Surface Temperature Data Set*

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Summary

The Meteorological Office Historical Sea Surface Temperature Data Set was created from the main marine data bank (Shearman 1983) to fulfil diverse needs in research into practical long-range forecasting and into climatic fluctuations observed since the mid-nineteenth century. This short paper provides an account of the origins, quality control, and applications of the data set.

1. Introduction

Ships' observations have been recorded since the mid-nineteenth century following the plan of Maury and Glaisher (British Meteorological Society 1852) to cover the whole world with observing stations. Many of these records were stored in archives of ships' logbooks until the 1960s when they were keyed into computerized data sets such as the 'TDF-11' data set created at the National Climatic Data Center, Asheville (North Carolina).

These data sets were supplemented by records for more recent years which have been exchanged internationally under the World Meteorological Organization's 'Resolution 35' carried in 1963. The main marine data bank of the Meteorological Office (Shearman 1983) was formed by combining these data sources, careful checks being made to remove any duplicate observations. The Meteorological Office Historical Sea Surface Temperature Data Set (MOHSST) has been created by extracting the sea surface temperature (SST) observations from the main marine data bank. Similar historical data sets have also been created for night-time and daytime marine air temperature. All these data sets contain monthly values for 5° latitude × 5° longitude areas, beginning at January 1854 and ending at December 1981.

2. Quality control

Observations of SST are beset by systematic biases, individual inaccuracies, and irregular distribution in space and time.

Systematic biases occur because of changes in instrumentation, siting, or procedures. The most notable example is the change from uninsulated bucket measurements (which were taken until the Second World War) to a mixture of engine-intake, hull-sensor, or insulated bucket readings. Bottomley *et al.* (n.d.) describe and incorporate compensation factors designed to take account of these changes. However, research into the effects of the systematic changes is not yet complete, and the data in MOHSST have not had any instrumental corrections made to them.

Individual inaccuracies were treated in the following way. First, no SSTs were included in the main marine data bank if they were outside the range −5 °C to 35 °C. Second, a provisional climatology with 1° latitude × 1° longitude and 5-day resolution was formed during the creation of MOHSST, and all SSTs deviating from this climatology by more than 6 °C were excluded. Third, after averaging over 1° latitude × 1° longitude areas for 5-day periods, the SSTs were converted into deviations from the provisional climatology and then subjected to a modified averaging process known as 'Winsorization'

* For greater details the reader is referred to the new Global Ocean Surface Temperature Atlas (Bottomley *et al.* n.d.) which is currently being produced jointly with the Massachusetts Institute of Technology.

(Afifi and Azen 1979). In this computation, which was made for each 5° latitude × 5° longitude area and month, values exceeding the top quartile were replaced by that quartile, and values below the bottom quartile were replaced by the bottom quartile. The adjusted set of values was then averaged. The resulting average is less influenced by outlying values than a straightforward average would be.

The problems of patchy distribution on the global scale have not entirely been overcome, and the geographical completeness of the climatology to be published by Bottomley *et al.* (n.d.) results only from the merging of a climatology based on MOHSST with an earlier climatology published by Alexander and Mobley (1976). This earlier climatology was complete only because the authors interpolated in the Southern Ocean between the southern limit of observed data and an assumed ice-edge temperature of −1.8 °C. Also, an earlier version of MOHSST was found to have major gaps in the data for the Pacific between 1961 and 1972; when creating the present version of MOHSST, these gaps were filled with values obtained from the US Navy Consolidated Data Set via the Massachusetts Institute of Technology (MIT).

The effects of irregular distribution of data within smaller areas were, however, allowed for in the quality control, firstly by comprehensive non-linear interpolation and smoothing of the provisional climatology within 10° latitude × 10° longitude areas, and secondly by working with 'anomalies' (deviations from climatology) on a geographical resolution of 1° latitude × 1° longitude and a time resolution of 5 days. The latter process avoids considerable biases; if an observation at 41° N, 18° W on 31 May 1856 was the only one in May 1856 in the area 40–45° N, 15–20° W, it would almost certainly be warmer than the average for that area and month even if it were colder than the local average for 31 May. The Winsorization described above, being applied to anomalies with respect to 1° latitude × 1° longitude areas and 5-day periods, produces a representative anomaly for the 5° latitude × 5° longitude area and month.

MOHSST is available as monthly 5° latitude × 5° longitude values computed by adding the quality-controlled anomalies to the smoothed climatology averaged over the area and month. Quality-control flags indicate where values failed extreme-value and scatter tests carried out during Winsorization and also where blank areas were supplemented with the data received from MIT.

3. Applications

MOHSST was created with the realization that it would find many uses. The major applications at present are:

(a) Input to statistical methods of long-range forecasting (Ratcliffe and Murray 1970, Palmer and Sun 1985, Folland and Woodcock 1986).

(b) Provision of boundary conditions for integrations of atmospheric general circulation models for

- (i) dynamical long-range forecasting (Mansfield 1986),
- (ii) assessment of the mechanisms underlying particular events, e.g. drought in sub-Saharan Africa, or hot summers in the United Kingdom (Folland *et al.* 1986), and
- (iii) investigation of the role of SST in climatic change.

(c) Formulation of relationships between SST and particular events, in order to gain understanding of the physical processes involved (Folland *et al.* 1986).

(d) Monitoring of climatic change, particularly with a view to assessing the effects of increasing concentrations of carbon dioxide in the atmosphere (Parker *et al.* 1986).

A dramatic recent application was to the study of drought in sub-Saharan Africa. The Meteorological Office Synoptic Climatology Branch's 11-layer global atmospheric general circulation model was integrated for 6 months of simulated time, firstly with SSTs corresponding to those observed from May to October 1950, and secondly with SSTs representing those observed in the same period of 1984. Fig. 1

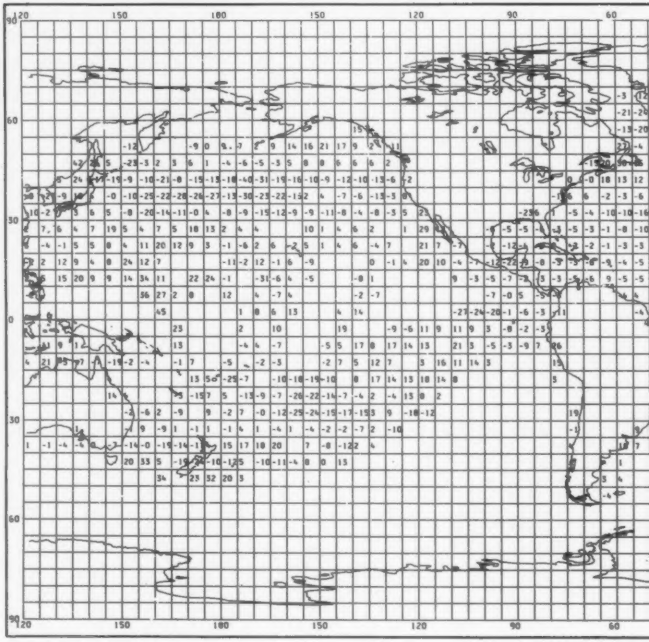
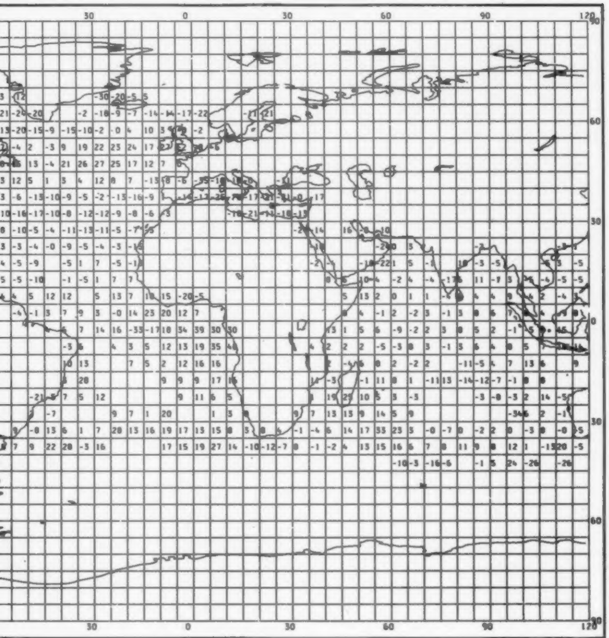


Figure 1. Sea surface temperature (tenths °



hs °C) August 1984 minus August 1950.

illustrates the SST differences between these years at the peak of the Sahel rainy season (August). The wettest year this century in the Sahel was 1950 and one of the driest was 1984. The model successfully reproduced the marked contrast of rainfall between the 2 years (Fig. 2). This study also served to elucidate some of the links in the chain of events between SST and Sahel rainfall, such as large-scale changes in atmospheric circulation and moisture convergence, and changes in soil moisture in the Sahel itself.

4. Future plans

The present version of MOHSST was created in 1983. Since then the main marine data bank has been improved by the addition of several data sources, which *inter alia* have covered the gap in Pacific Ocean data in 1961–72. The data bank has also been updated. Plans are being made, therefore, to produce a new version of MOHSST from the main marine data bank. The quality-control procedures will be similar, but separate daytime and night-time data sets will be created so that time-of-observation biases can be estimated. These are important for assessing the SSTs obtained from polar-orbiting satellites which sometimes have nearly fixed transit times. New daytime and night-time data sets of marine air temperature will also be created.

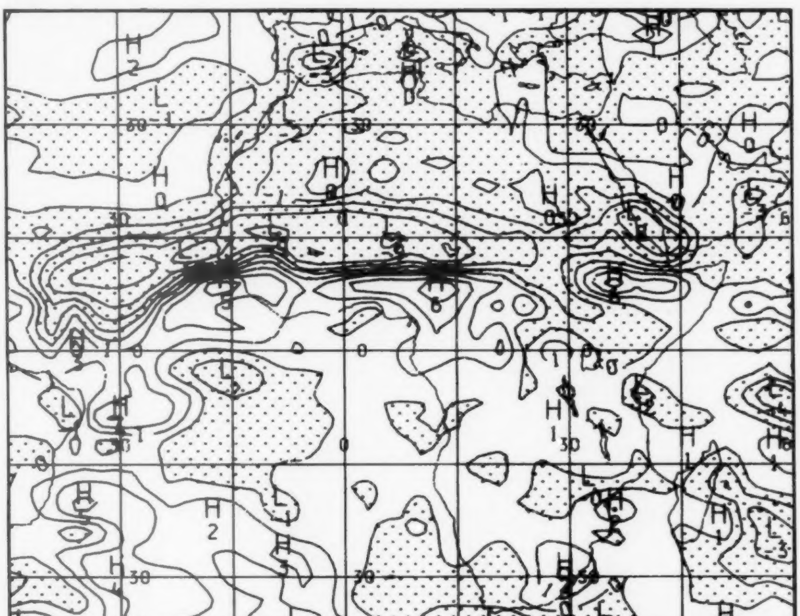


Figure 2. Model rainfall (mm d^{-1}) August 1984 minus August 1950. Stippled areas are negative (1984 drier).

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Dealing with winter chaos*

R.D. Hunt

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Summary

Transport chaos in winter is costly for industry and the country. Features of the British weather are briefly discussed together with some of the developments taking place in the Meteorological Office and elsewhere to deal with the problem.

1. Introduction

There are a number of features of winter weather in the United Kingdom which create special problems. Two of the more significant are:

(a) The weather is rarely severe for very long; the number of days with severe weather is a very small percentage of the total number of winter days. This makes the allocation of financial resources to deal with the problem very difficult.

(b) The weather situations which bring chaos to the country are notoriously difficult to forecast in detail because they are often very marginal. While forecasting technology has increased to such an extent nowadays that the onset of a cold spell can be predicted several days in advance (and indeed much publicity was given to the correct forecast of the massive change to severe weather in January 1987), trying to pinpoint where snow is going to be heaviest, just how much will fall, whether or not it will lie, etc. is very difficult.

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* A lecture presented to the Institution of Civil Engineers' Conference, 'Winter chaos — can we buy our way out of it?', London, 31 March 1987.

In countries with a more continental type of climate, temperatures stay below freezing for long periods of time, frost is often a near certainty and any precipitation which falls can be expected to be snow, usually of the dry, powdery kind. In the British Isles there is often doubt as to whether the temperature will fall below freezing point at night because of difficulties in assessing cloud amount or wind speed, both of which have an affect on temperature, while the situations which lead to widespread heavy snow usually involve the coming together of cold and mild air; there is, therefore, uncertainty about whether to expect snow or rain. What snow does fall is often wet and sticky.

Snow usually occurs in the United Kingdom when the temperature is close to freezing point. Even when snowfall can be predicted with a high degree of confidence, there is still the problem of deciding whether it will lie at all levels, or just on higher ground. A drop in temperature of just 1 °C can make all the difference between a wet day and a day of winter travel chaos.

There is no question that accurate forecasts in winter are of great financial benefit to whole sections of industry as well as to the authorities responsible for road and rail transport. The developments outlined below indicate how the Meteorological Office is trying to improve its forecasts of ice and snow and also the liaison with the transport authorities in order to reduce the effects, and costs, of winter chaos.

2. Computer forecast models

The Meteorological Office has developed sophisticated computer models for forecasting weather around the world for a week ahead. One model is used for producing detailed weather forecasts in the British Isles for the following 36 hours. It has a resolution of about 75 km in the vicinity of the United Kingdom and is capable of predicting not only the pressure patterns and fronts, familiar from television and newspapers, but also where and how much rain is likely to fall. By calculating temperatures through the atmosphere, it is also able to indicate probabilities of precipitation falling as snow rather than rain.

The results from these models have been very impressive and are probably the main reason why the general public's perception of weather forecast accuracy has risen so much in the last few years. Nevertheless, the computer is unable to produce all the answers; there is still a great deal of human input necessary both in interpreting the information and in assessing up-to-date observations.

Although the resolution of 75 km is good for many purposes, the weather can change over much smaller distances than that. Hills and coasts affect the weather over small areas, while bands of rain and showers often get organized on a small scale. Because of this, a new model has been developed with a resolution of only 15 km to cover the whole British Isles. This should become fully operational soon and will lead to more accuracy in forecasts a day ahead, including forecasts of snowfall and frost. Initial results from the development phase have been encouraging.

3. Weather radar

Weather radar is now used by forecasters to see not only where rain or snow is falling but also how much. By looking at a time sequence of displays over the most recent few hours, judgements can be made about which areas are most likely to receive snow in the next few hours.

The radars are quite expensive and their range is fairly limited. Nevertheless, it has been possible for the Meteorological Office, in conjunction with the Water Authorities and other interested parties, to cover large areas of England, Wales and Northern Ireland with a radar network. Information from this network has already been an invaluable aid. During the severe weather in January 1987, for instance, forecasters were able to see just how far snow showers coming across the coast were penetrating inland and to assess what track they were taking. It was, therefore, possible to indicate which areas were likely to be most affected. (Unfortunately Kent is on the edge of the radar picture and so the data there were of less use, particularly as the snow was coming in from the North Sea.) Similarly, in conditions of more

Photograph by courtesy of A.J. Byrne

general precipitation it was possible to assess the speed of the spread of snow and to indicate where it was heaviest. Plans are already well advanced to expand the network further, in eastern England, in the south-west and into Scotland.

4. Road surface temperature forecasting

For many years the Meteorological Office has been able to give warnings of night frost in a general sense. But the increasing cost of salting roads has led to the development of more sophisticated techniques for forecasting temperatures actually on road surfaces. Models have been developed, both in the Meteorological Office* and at the University of Birmingham, which can forecast road temperatures for the overnight period thereby not only showing whether sub-freezing temperatures are expected but also giving the time at which the temperature is expected to fall below freezing and the minimum temperature.

In order for the models to produce results, detailed forecasts of cloud amounts, air temperature and wind speed are required, so improvements in general forecasting mentioned earlier are important. The output from this, or any other model, is only as good as the data put into it.

Also, for the model forecasts to be of maximum benefit, they should be for specific locations at which there are road sensors installed. If road temperature information is available in real time, the forecaster can update his forecast as necessary — he can also verify the output. This increases the confidence in the

* For details see P.J. Rayer; The Meteorological Office forecast road surface temperature model, *Meteorol Mag*, **116**, 1987, 180–191.

forecasts and assists with further development of the model. Road sensors can be placed at locations of most benefit to the local authorities, making use of thermal mapping techniques to select cold spots and to relate temperatures at one site to a much wider area.

5. Practical assistance to transport maintenance and planning

The winter of 1986/87 has seen the biggest growth ever in the amount and quality of services provided by the Meteorological Office to assist maintenance of the road and rail network. A new service, called Open Road, was launched in the autumn of 1986 and many local authorities have taken advantage of it.

Open Road has combined many of the features described above to produce a forecast service as accurate as possible within the current state of the science. Road temperature forecasts are provided routinely to the authorities, together with detailed forecasts in pictorial and text form and copies of the weather radar output with comments added by forecasters. Apart from forecasts for the coming night, planning outlooks to 5 days are provided as well as a summary of the previous night's weather. Similar services have been arranged with many of the British Rail regions and also the London Underground.

One of the attractions of Open Road is that it has led to an increase in the liaison between the local authorities and the forecasters. The improved contact, as well as the better product, has given rise to a much higher level of confidence in the forecasts. It should also be stressed that far more information is coming back from the authorities as well. Latest reports about conditions around the counties are vital to the forecaster who, even with satellite pictures, radar, and computer products, still needs more knowledge about what is actually happening on the ground.

Apart from giving information on expected weather conditions, the Meteorological Office has been able to give considerable advice on the positioning of new roads and motorways in order to minimize the occurrence of fog and high winds and to maximize the use of warning signs. This is very much a growth area and one which should lead to more efficient use of financial resources.

6. Concluding remarks

Severe weather is nearly always going to occur at some stage in a British winter. Sometimes conditions may be so severe that transport chaos would occur even if forecasts were perfect; indeed January 1987 may well have seen extreme conditions, in parts of Kent especially, when this would have applied. But, generally speaking, an accurate forecast a few days ahead of the possibility of wintry weather combined with detailed forecasts out to about 1 day ahead can be of great practical assistance.

Because of the developments described above, and others not mentioned here, forecasts are now far more accurate than they used to be. This enables a much more comprehensive service to be provided. Obviously the amount of work and technology involved is much larger than hitherto and this is reflected in the cost. Nevertheless, costs are very low compared to the total winter maintenance budget.

As in so many other areas, it is necessary to spend some money in order to save considerably more. Winter chaos can be alleviated by taking advantage of improved weather services, but not eliminated.

Review

Contemporary climatology, by A. Henderson-Sellers and P.J. Robinson. 157 mm × 234 mm, pp. xvi + 439, *illus.* Harlow, Longman, 1986. Price £12.95.

For most subjects, there are now many books to which one can refer for up-to-date information. Amongst the meteorological titles, *Contemporary climatology* is worthy of consideration. This joint

offering by A. Henderson-Sellers and P.J. Robinson, of the Geography Departments at the University of Liverpool and North Carolina respectively, was written in 1984 and is aimed at undergraduates without a strong mathematical background. It generally succeeds, although the level of complexity varies greatly, thus making sections of it potentially useful to the interested layman. Equations are few and far between, but the text carries the reader lucidly through a variety of concepts, supplemented by many clear illustrations and black and white photographs together with a useful glossary of climatological and meteorological terms, an appendix of SI units and a comprehensive index.

The book can be effectively divided into two parts. Part 1 (chapters 1–3) deals with the science of climatology from its early beginnings through its unhappy recent past ('...climatologists were the halt and the lame...') to its current place as a subject with important contributions to make. Basic ideas that govern the behaviour of the climatic system are introduced and these include a discussion of the radiation budget of the earth and the importance of the latitudinal imbalance in absorbed and emitted radiation, the hydrological cycle and the effects of water vapour in the atmosphere, cloud formation, evaporation and global precipitation distribution. The difficulties encountered in obtaining accurate measurements of precipitation and evaporation are brought to the reader's attention but a consideration of soil moisture deficit and its seasonal variation, a parameter that can affect precipitation amounts, is missing.

This discussion of physical climatology then gives way to a study of dynamical aspects in Part 2 (chapters 4–7). The general circulation and global climate are introduced and ideas on the causes of the circulation, Coriolis force, barotropic and baroclinic conditions, Rossby waves, etc. are stated in a simple but effective fashion. A change of scale leads to a consideration of regional climates and the methods used to classify them. The Köppen system is widely used and this is clearly explained. Tropical, mid-latitude and polar climates are examined and results from a simulation model, used to quantify precipitation changes resulting from deforestation of the Amazon basin and its conversion to savannah grassland, are included. Our attention is then turned to local climates, where we are introduced to the influence of topography as a modifying force, pollution, the human response to climate and the effect of human activities on climate. Very scant attention has been given to what is perhaps one of the more important effects of increasing urbanization, namely its tendency to increase local precipitation totals, a finding that arose out of the METROMEX program. A particularly interesting diagram is included of the locally named winds of the Mediterranean basin, approximately 50 of them. In rounding off the book, the authors look to the future and ponder the issue of climatic change. The possible causes of such a change are discussed, the models that simulate climates and the problems associated with these models are considered, and the evidence to suggest that climatic change has taken place in the past is presented. Finally, the carbon dioxide problem is given considerable space.

Both authors are well known and have collaborated to produce a comprehensive and comparatively error-free book, which will no doubt find its way on to many bookshelves. Two points of criticism; firstly, the narrative, although clear, fails in my opinion to convey the great excitement which can characterize a study of the earth's climatic systems, and secondly, why was the book produced at all? Much of what is between the covers of this work has appeared in other books — do we really need this degree of repetition?

Nevertheless, a great deal of care has been taken by the authors in its preparation and for that reason, it deserves to be a success.

M.S. Shawyer

Books received

The listing of books under this heading does not preclude a review in the Meteorological Magazine at a later date.

Landolt-Börnstein: numerical data and functional relationships in science and technology, V/4a, edited by G. Fischer (Berlin, Heidelberg, New York, London, Paris, Tokyo, Springer-Verlag, 1987. DM 1220.00) contains meteorological data, mainly in tabular and diagrammatic form, on the physics of the atmosphere. This subvolume covers the thermodynamic structure and dynamics of the global atmosphere.

Remote sensing digital image analysis, by J.A. Richards (Berlin, Heidelberg, New York, London, Paris, Tokyo, Springer-Verlag, 1986. DM 138.00) includes an overview of remote-sensing data sources and characteristics. It seeks to draw together the range of digital image-processing procedures into a single treatment.

A glossary of computing terms, edited by the British Computer Society Schools Committee Glossary Working Party (Cambridge University Press, 1987. £1.95, US $3.95) contains an explanation of over 800 computing terms divided into 15 sections, but with a complete index also. All aspects of computing are included.

Monsoons, edited by J.S. Fein and P.L. Stephens (New York, Chichester, Brisbane, Toronto, Singapore, John Wiley and Sons, 1987. £71.75) consists of 19 chapters on many aspects of the subject and takes a broad multi-disciplinary look at these variable natural phenomena.

Antarctic science, edited by D.W.H. Walton (Cambridge University Press, 1987. £25.00, US $39.50) reviews the major international developments in Antarctic science from its earliest beginnings, and attempts to forecast future developments in the area. Biology, the earth sciences and atmospheric science are examined, with emphasis on the achievements of the past 25 years.

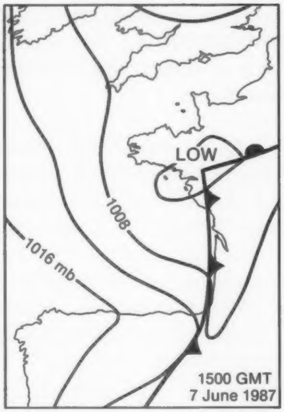
Atmospheres and ionospheres of the outer planets and their satellites, by S.K. Atreya (Berlin, Heidelberg, New York, London, Paris, Tokyo, Springer-Verlag, 1986. DM 148.00) presents a theoretical discussion of the subject's relevant physical and chemical processes. Observational data, methods of their retrieval and the shortcomings of theoretical models are discussed with a view to further research.

Climate and plant distribution, by F.I. Woodward (Cambridge University Press, 1987. £8.95 (paperback), £22.50 (hardback)) examines the thesis that climate exerts the dominant control on the distribution of vegetation types. Firstly the distribution of species in relation to climate over different scales of time and place world-wide is considered, and secondly the approaches to explaining observed correlations between plant distribution and climate, and the mechanisms of physiological and biochemical control, are investigated.

Satellite photographs — 7 June 1987 at 1442 GMT

The infra-red (left) and visible (right) images show cloud over France and the Bay of Biscay associated with an eastward-moving low, and cold front. Considerable dense 'lumpy' cloud in both images indicates deep convection, especially along the cold front where an organized band of anvils is seen. Within the warm sector of the low, widespread layered cloud is present which over the mountains of northern Spain is broken into lee waves.

Detailed re-analysis of all the available surface observations near the cold front as it crossed western France indicated that: the front lay close to the forward edge of the cloud band (possibly along the narrow band of bright cloud in the south-east corner of the Bay of Biscay, as seen on the visible image); it was accompanied by a squall line some 300 km in length where winds suddenly changed from 5–10 kn southerly to 25–35 kn westerly, with 50–60 kn gusts; the front moved eastwards at 36 kn; surface pressure rose by about 3 mb; and surface temperatures fell by as much as 12 °C (from 26 °C to 14 °C at Bordeaux). The suddenness of the arrival of the squall line led to the deaths of at least eight people, mostly through drowning.



Photographs by courtesy of University of Dundee

Meteorological Magazine

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Meteorological Magazine

August 1987

Editor: R.W. Riddaway

Vol. 116
No. 1381

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ISBN 0 11 727972 2 ISSN 0026–1149
